ISSN 0742-0463, Journal of Volcanology and Seismology, 2008, Vol. 2, No. 5, pp. 356–363. © Pleiades Publishing, Ltd., 2008.
Original Russian Text © A.P. Maksimov, 2008, published in Vulkanologiya i Seismologiya, 2008, No. 5, pp. 60–68.

# A Physicochemical Model for Deep Degassing of Water-Rich Magma

**A. P. Maksimov**

*Institute of Volcanology and Seismology, Far East Division, Russian Academy of Sciences, Petropavlovsk-Kamchatskii, 683006 Russia*

Received June 14, 2007

**Abstract**—Two powerful eruptions of Quizapu vent on Cerro Azul Volcano, Chile are used as examples to discuss the problem of effusive eruptions of magmas having high preeruptive volatile concentrations. A physicochemical mechanism is proposed for magma degassing, with the volatiles being lost before coming to the surface. The model is based on the interaction of magmas residing in chambers at different depths and on the difference between the solubility of water in the melt and the water equilibrium concentration in a magma body having a considerable vertical extent. The shallower chamber can accumulate the volatiles released from the magma that is supplied from the deeper chamber. An explanation is provided of the dramatic differences in the character of the 1846–1847 and 1932 eruptions, which had identical chemical–petrographic magma compositions.

**DOI:** 10.1134/S0742046308050059

## INTRODUCTION

The relationship between andesitic and more silicic magmas and the increased content of volatiles in them is well-known. One manifestation of this relationship is the highly explosive character of volcanic eruptions that discharge such magmas; the ejecta frequently consist of ash and pumice tephra, and pyroclastic flows. The Plinian character of these eruptions is a natural consequence of the high concentrations of volatiles in the magmas before the eruption, i.e., in the feeding chambers.

However, there are also examples of purely effusive discharges of silicic magmas. Dacitic flows have been described for Krasheninnikova Volcano, Kamchatka [6]. These flows are highly fluid (they can be as long as 13 km on a terrain of little slope, while their thickness is a few meters only) and are nearly devoid of accompanying tephra deposits.

The problem of nonexplosive silicic volcanism was discussed by Eichelberger et al. [10]; these authors note the paradoxical difference in the concentrations of volatiles for comagmatic tephra eruptions and extrusive domes, with these rocks being petrologically similar. Two mechanisms have been suggested for the transition from explosive to extrusive and effusive volcanism. According to one of these, the transition may be caused by a preeruptive gradient of the volatiles in the magma residing in the chamber. In this case the change in the eruption character reflects the ascent of magma that was originally progressively more depleted in volatiles. According to the other mechanism, the transition to nonexplosive eruption results from a redistribution of the volatiles in magma that was originally homogeneous in volatile content during the ascent and associated degassing.

Special interest attaches to the cases of effusive discharges of silicic magmas that contain hydrous minerals. The presence of such minerals indicates high preeruptive concentrations of water in the melt. That last circumstance must produce high explosivity, hence the effusive character of such eruptions calls for a special explanation.One example of such an eruption is that of 1846–1847 on Quizapu crater, Chile. The short-lived explosive phase of this eruption manifested itself in individual initial explosions followed by a long-continued effusive discharge of great volumes of hornblende dacite lava [12]. The insignificant role of the explosive phase during the 1846–1847 Quizapu eruption makes it impossible to explain the degassing of so large a volume of water-saturated magma by mechanisms suggested in [10]. This problem was considered by W. Hildreth and R. Drake [12], but they have not managed to find a convincing solution. The materials quoted in that reference are used in the present study to propose a viable model of preeruptive degassing that can explain the effusive character of discharge for magmas that were originally rich in volatiles.

## A BRIEF HISTORY OF VOLCANIC ACTIVITY AT QUIZAPU CONE

Quizapu is a small parasitic cone on the flank of Cerro Azul Volcano, Chile. Below we present some information on Cerro Azul Volcano and characterize the eruptions of Quizapu and the associated ejecta; a detailed account can be found in [12].

Cerro Azul Volcano with the Quizapu cone is situated within a complex volcanic group. Besides two major stratovolcanoes, this volcanic group includes numerous Holocene and Late Pleistocene eruption centers producing cinder and lava ranging from basic to rhyodacitic composition. During historical times, however (since the Europeans first came to settle there in the early 16th century), neither eruptive nor fumarole activity had been recorded in the area until 1846. The volcanic group stands on a thick old granodiorite intrusive body. The Cerro Azul edifice is composed of a wide range of rocks from basalts to pyroxene dacites (51–69% $SiO_2$). No amphibole was found in this volcano's rocks prior to the first Quizapu eruption in 1846–1847. The youngest (before the Quizapu eruption) lava flow is composed of low-crystalline hornblende-less rhyodacites (69.2% $SiO_2$).

The generation of Quizapu cone was associated with two giant eruptions. The first occurred in 1846–1847. Except for a very short initial phase, the eruption was nearly entirely effusive, discharging 5 $km^3$ of dacitic hornblende lava. Subsequently, the volcano was in a state of repose for several decades.

Eruptive activity occurred between 1907 and 1932, at first weak activity and then more intensive activity, featuring ash columns taller than 4 km and nighttime luminescence. It was during that period that the Quizapu cone proper was formed.

A powerful explosive eruption occurred on Quizapu in April 1932, discharging 9.5 $km^3$ of pumice. According to varying estimates, the eruptive column varied between 10 and 30 km in height. The Plinian phase lasted about 18 hours. This eruptive activity nearly ceased after 7–10 days. Two weeks following the Plinian activity, fumarole activity began 6–7 km north of Quizapu on the flank of Descabezado Grande stratovolcano. This was the first manifestation of activity on the volcano during historical times. As well, a new crater 600 m wide formed on the north flank of Descabezado Grande, showing episodic weak to moderate explosive activity until 1933. The result was the production of a layer of resurgent deposits without juvenile components. No eruptions have occurred since 1933.

**The 1846 and 1932 ejecta.** The 1846 and 1932 Quizapu eruptions discharged ~5 and 4 $km^3$ of hornblende dacite magma, respectively (converted to solid rock). Until teh Hudson Volcano erupted in 1991, these two had been the largest historical eruptions in South America during historical times.

The 1846–1847 ejecta are exclusively composed of lavas, no ash has been found. The lavas are hornblende–hypersthene–plagioclase dacites with little pyroxene and 67–68%$SiO_2$. The lava flows contain abundant magma inclusions of basic to intermediate composition with olivine and plagioclase phenocrysts, usually having chilled boundaries. The 1846 lavas were the first on Serro Azul to contain hornblende.

Over 95% (possibly 98%) of the 1932 rocks are the same (mineralogically and chemically, ~68% $SiO_2$) dacites, but the 1932 eruption began and terminated with discharges of black olivine-bearing cinder. The cinder is in the form of andesitic bombs and banded dacite–andesite pumice. A thin layer of andesitic and banded pumice can also be found among Plinian dacites. The amount of silica in analyses of the original cinder is in the range 52–63% $SiO_2$. Another juvenile component is foamed rhyodacitic pumice with 69–70% $SiO_2$; this amounts to less than 0.5% of all deposits, but is found everywhere. Rocks of the granodiorite intrusive phase (occasionally subjected to partial fritting) are present among the resurgent fragments, in addition to the 1846–1847 lavas and rare samples of the older Serro Azul rocks.

The dacites of these eruptions on the cone fit into a narrow range of acidity, ~67–68% $SiO_2$. The 1932 eruption typically exhibits an extremely wide range of ejecta, from basalts to rhyodacites (52–70% $SiO_2$), with the acid dacites strongly predominating.The wide composition range of this eruption provides evidence of interaction between the basic (basaltic) and silicic (dacitic) magmas, with the result being their physical and (in part) chemical mixing. The basic and intermediate magma inclusions in the 1846–1847 lavas also point to an interaction of dacitic and more basic magmas. This is confirmed by a linear trend in the variations of major oxides in relation to silica content for the ejecta of the two eruptions [12]. No hornblende is usually found in those 1932 rocks with < 63% $SiO_2$.

The dacites discharged by the two eruptions are nearly identical in mineralogic, chemical, geochemical, and isotope composition. The respective dacites have almost indistinguishable compositions and amounts of phenocrysts (15–19%). The hornblende in the 1932 dacites and in chilled glassy parts of the 1846–1847 lavas does not have opacite reaction rims. The preeruptive temperatures and the reduction–oxidation mode of the magmas calculated from the compositions of coexistent titanomagnetite–ilmenite couples are identical for the 1846–1847 lavas and the 1932 pumice. Lower temperature and oxygen fugacity were obtained for the andesitic cinder discharged during the initial phase of the 1932 eruption. The only significant difference between the 1846–1847 and 1932 dacites is observed in the isotope hydrogen ratios for the respective hornblendes. Both of these dacitic magmas were certainly discharged from a common reservoir. However, the first eruption was effusive, while the second an effusive, Plinian-type one.

## FORMULATING THE "QUIZAPU ENIGMA" PROBLEM

Among the features of Quizapu volcanic activity described above the following stand out:

(1) the 1846–1847 and 1932 eruptions are sharply different in character, even though the chemistry and mineralogy of the respective ejecta, as well as the discharged volumes, are similar;

(2) the almost effusive mode of eruption in 1846-1847, as contrasted with the high acidity of the rocks and the presence of phenocrysts of unaltered hornblende in these.

The last feature suggests an interesting physicochemical problem stated in [12]: "How could the 5 km$^3$ of the 1846–1847 dacitic hornblende magma degass in a passive manner?"

Before we proceed to set forth our solution to the problem, we are going to discuss its physicochemical aspects in more detail. Such an analysis is reported in the paper referred to above and is given below with our additions.

Using the experimental data to be found in [15–17], the authors remarked that the presence of fresh (equilibrium) hornblende phenocrysts is evidence of a high water content in the deep melts, at least 4–5 wt %. At the same time, the content of water in the 1846–1847 dacites is a mere 0.1–0.3%. It follows that at least 500 million tons of water must have been released from the magma, corresponding to 0.5 km$^3$ of liquid water. The enigma is not only why the eruption was effusive rather than explosive, but also where the water has gone, since neither intensive fumarole activity nor hot springs had been noticed prior to the beginning of the 1907 explosive activity.

The effusive character of the first eruption thus implies preeruptive degassing of dacitic magma. The inference that the dacitic magma of the 1846–1847 Quizapu eruption largely degassed in a deep environment was also based on the analysis of hydrogen isotopes in the ejecta of the two eruptions [12].

In contrast to this, the magma of the 1932 dacites experienced a sharp near-surface and surface degassing, which manifested itself in the Plinian character of the eruption. The amount of water remaining in these rocks is appreciably higher than that in the analogous 1846–1847 dacites (0.74% on the average). That is, in both of these cases, even if the melts had been only partially degassed, the original amount of water was more than sufficient to produce a catastrophic explosive eruption, but the volatiles were lost in the first of these cases in an entirely different manner, more likely at depth.

While allotting the leading role in the degassing of dacitic magma to decompression, Hildreth and Drake [12] admit some influence on the process on the part of heating and subsequent convection due to hotter mafic magma coming to the acidic chamber from below. There are data to substatiate this assumption in the shape of chilled mafic inclusions scattered in the 1846–1847 dacites. However, the factual material opposes any appreciable part being played by heating during magma degassing. In the first place, the preeruptive crystallization temperatures for the 1846–1847 dacites practically do not exceed those for the 1932 dacites, and even cover lower values. Secondly, the dacites of these eruptions contain identical sets of phenocrysts and the same percentage of these. It should be noted that the presence of hornblende among them, which has the highest temperature of stability in dacite, about 950°C [14, 15, 18], puts a serious constraint on a significant heating of the magma. Also, heating of magma in the presence of phenocrysts would melt them, hence would cause a corresponding increase in the fraction of melt, and can only produce an additional solution of water (volatiles). Taken by itself, the influence of temperature on the solubility of water in melts is small. Cooling and the associated crystallization might have been more significant. However, that mechanism is to be rejected as well by virtue of the above argument. Obviously, the temperature regimes for the dacitic magmas of both eruptions were analogous, and temperature variations cannot account for the differences in the eruption character, while the degassing was caused by decompression of the magma during its ascent from the chamber.

The hypothesis that most of the water was released from the 1846–1847 dacitic magma during a long-continued slow ascent [12] is in poor agreement with experimental evidence for the stability of hornblende and the rate of its breakdown at low pressures [15–17]. An aqueous dacitic magma at a temperature of 860°C must necessarily be degassed during the ascent to depths shallower than ~4 km, while amphibole must break down. Since there are no opacite reaction rims on hornblende in the chilled parts of the 1846–1847 lava, it follows that magma will not take longer than a few days to rise to the surface from the region of amphibole stability, as can be inferred from experiments on the rate of opacite reaction rim generation on hornblende [16]; hence the time of separation for ~4.5% of water must probably be within the same time interval. Even if the hornblende breakdown under natural environments is retarded by additional factors compared with the experimental conditions, the attempt at explaining preeruptive degassing by a slow ascent rate of the first portion of the 1846 magma is unacceptable, because it adds nothing essentially novel.

Another feature in the Quizapu activity is to be noted, viz., a long-continued, moderate, explosive activity from 1907 to 1931. Hildreth and Drake [12] regard the volcano's activity during that period as largely phreatic or occasionally fumarole, and the associated ash as mostly resurgent. However, their description convinces one that this is far from being the case. The presence of black cinder layers in the sections of the crater, active luminescence, long-continued persistent Vulcanian-type activity with ash columns taller than 4 km, all testify to a preeminently magmatic Vulcanoan–Strombolian type of activity. As well, this period immediately preceded one of the major Plinian-type eruptions in the Andes during the 20th century.

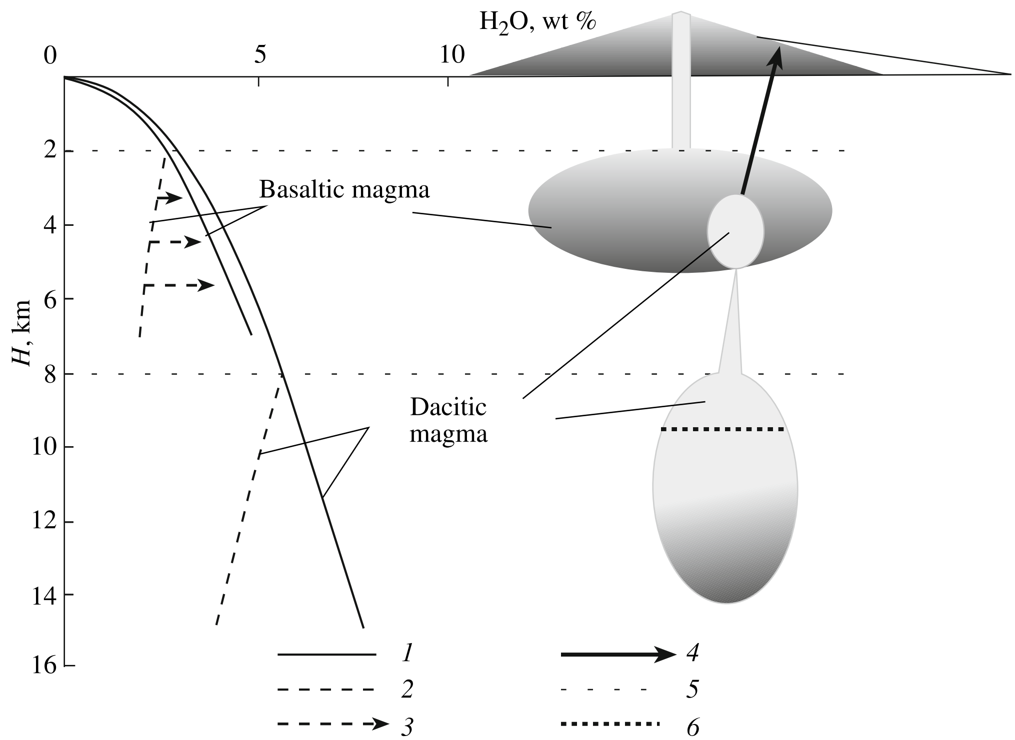


**Fig. 1.** A diagrammatic illustration of the model of deep magma degassing. Left in this depth (*H*, km) versus water concentration ($H_2O$, wt %) curve are: (*1*) curves of water solubility in a melt of basalt and dacite (calculated from the solubility model after [8, 13]), (*2*) approximate positions of the curves of equilibrium distribution of water against depth in the chamber, for basaltic and dacitic melts, (*3*) direction of evolution for water concentration in the shallow chamber during the passage of degassing magma coming from the deeper chamber. Right is a schematic representation of the feeding system for Quizapu in 1846–1847, (*4*) direction of motion of dacitic magma toward the surface, (*5*) schematic positions of the tops of chambers over depth, (*6*) assumed lower boundary of dacitic magma in the deeper chamber.

## THE MODEL OF DEEP DEGASSING

Apart from answering the main question of a mechanism of preeruptive degassing, the model must account for other facts as well:

(1) the wide compositional range of the 1932 juvenile ejecta;

(2) the absence of amphibole from the products of Serro Azul prehistoric activity and from the more basic varieties of the 1932 ejecta;

(3) a period of long-continued moderate explosive activity over a few decades following the first eruption on Quizapu;

(4) a good state of preservation of hornblende in the discharges of Plinian eruptions on Quizapu.

In our opinion, the “Quizapu enigma” can only be explained by the existence of two magma chambers at different depths beneath the volcano (Fig. 1). The shallower chamber must lie at depths of a few kilometers, where amphibole is unstable in melts at any temperature and water concentration. This chamber supplied magma to Cerro Azul during the pre-Quizapu period, considering that the rocks of the volcanic edifice, among them the most recent rhyodacitic flow, contain no amphibole. Since Cerro Azul itself is composed of a wide range of rocks from basalt to rhyodacite, the chamber must have been zonal in composition. The lower and intermediate horizons of the chamber seem to have been of basaltic and basaltic andesite composition prior to the Quizapu eruptions, while the upper horizons must have been andesitic, judging from the compositions found for the 1846–1847 inclusions and the 1932 cinder. As mentioned above, amphibole is also absent from the inclusions and cinder. It should be emphasized that the absence of hornblende is due to low pressures, not to low concentrations of water in the magma. Even when the concentration of water is low, which is the usual case for basic magmas, enough water is accumulated in the magma, if crystallization differentiation proceeds as far as dacite and rhyodacite, but at low pressures most of the water cannot dissolve in the melt and becomes liquid.

The deeper chamber supplied dacite for the 1846–1847 and 1932 eruptions. The minimum depth at which dacitic magma can have occurred before these eruptions is determined by the lowest pressure at which hornblende can crystallize from the melt. That lowest pressure is determined by the point of intersection between the curve of amphibole stability and the solidus curve. According to several experimental studies

[3, 14–18], amphibole is stable in andesitic to rhyolite melts at pressures of more than ~1 ± 0.5 kbars, i.e., at depths greater than 2–4.5 km. However, hornblende crystallizes in natural magmas at temperatures considerably above the solidus, i.e., the amount of phenocrysts in amphibole-bearing rocks does not generally exceed 30–40%. The percentage of phenocrysts in the Quizapu dacites does not exceed 20%. Accordingly, the lowest pressure at which amphibole-bearing rocks were crystallizing is increased to reach 2–3 kbars (corresponding to depths of ~6–9 km). The lower chamber is likely to have also undergone a deep crystallization under closed environments, with water accumulating in the melt during cooling. This chamber evolution led to the generation of hornblende dacitic magma in the upper horizons of the chamber and of rhyodacitic magma in the chamber top itself, because more basic magmas were situated below. A physicochemical model of such an evolution is described in [3]. The long-continued closure of this chamber and the deep differentiation of the melt were favored by the chamber residing at great depths, as well as by the long repose period of the volcano, while the supply of mantle melts to the feeding system of the volcano is likely to have ceased long ago. This hypothesis is supported both by a wide occurrence of andesite, dacite, and rhyolite among the Cerro Azul ejecta and by the long repose period. The sufficiently deep position of the dacitic magma prior to the eruption is geologically supported by the fact that the 1932 dacites contain xenoliths of the thick ancient granodiorite massif underlying the entire volcanic group.

Magmatic activity was resumed after this long repose period, seemingly when the pressure of volatiles in the lower chamber exceeded the strength of the overlying rock. The volatile-oversaturated, dacitic magma rising from the upper horizons of the deeper chamber came into the shallower chamber, which was mostly filled with more basic magma.

Three circumstances are to be noted. First, because the silicic and the more basic magma have very different viscosities, they could not effectively mix during a comparatively short period of time when the dacitic magma was rising to the surface. What did take place was a limited physical mixing of the basic magma into the silicic accompanied by some partial chemical mixing; the result was to produce banded pumice and cinder of basic and intermediate composition. In that case the silicic magma must, because of its lower density, float through the basaltic magma. Secondly, because the silicic magma was oversaturated with volatiles, it was continuously undergoing degassing during the decompression as it was moving upward. This process of the dacitic magma ascent was gradually assuming an avalanche-like character at the shallow upper chamber, because effects of the volatile-losing silicic magma were increasing in volume and decreasing in density and became more important. This accelerated flotation. The magma was not moving along a narrow dike or a channel for much of the way to the surface, but was rising in the comparatively free space of a low-viscosity basaltic magma; this may have been an extra factor to help amphibole quickly reach the surface and accordingly to keep it in an excellent state of preservation. Thirdly, the magma in the upper chamber, except for its top, was not saturated with volatiles, which follows from the requirement of thermodynamic equilibrium. This last circumstance forms the physicochemical basis for the model here proposed.

It has been shown by several researchers [2, 4, 7] that the requirement of thermodynamic equilibrium for a vertically extended magma body causes the concentration of volatiles, including water (the primary agent among them), to decrease with depth. The actual distribution of volatiles in a chamber also depends on many other factors, in particular, the presence of convection, the rate of convection, the supply of material from below, the history of the chamber dynamics, the temperature distribution, etc. Nevertheless, to provide for the long-term stable existence of the chamber, the concentration of volatiles must decrease with depth, hence must be significantly lower at lower horizons in the chamber than can be deduced from the solubility curve. This is illustrated schematically in Fig. 1, which shows the solubility curves for water in basaltic and dacitic melts as calculated from the Burnham model [8, 13], and approximate positions of the water equilibrium distribution curves along the depth of the magma column. An exact computation of equilibrium distribution curves is impeded by the lack of necessary data on structural melt units and their thermodynamic properties.

Considering that the volcano had long been in repose, crystallization must have continually been occurring at the top of the shallow chamber. Accordingly, water was accumulated in the magma of the upper near-contact region of the chamber and the melt became volatile-saturated. With increasing depth the concentration of water decreased (in accordance with the requirement of thermodynamic equilibrium and because of the low diffusion rate for water in the melt), so that the difference between the solubility and the equilibrium concentration of water was progressively increasing (Fig. 1).

The dacitic magma was reaching saturation or near it owing to the same factors, such as being in the upper portion of the deeper chamber. This is corroborated by the presence of near-liquidus amphibole phenocrysts. Because the solubility of water in the melt increases with depth, it follows that the dacitic magma was originally much richer in water than the magma in the shallow chamber, especially in its lower horizons. Dacitic magma undergoes active degassing as it is rising to the surface. When it penetrates in the lower horizons of the shallow chamber, which are not saturated with water, further movement of the silicic magma through basic magma must inevitably be accompanied by the volatiles being dissolved in the basic magma, until the latter reaches saturation. Figure 1 illustrates the process diagrammatically by arrows pointing from the equilibrium distribution curve toward the saturation curve. The shallow chamber can thus act as a kind of "trap" for released volatiles.

The process outlined above is applicable to the first historical eruption of 1846–1847, that is, to the first passage of a portion of silicic magma through the shallow chamber. The dacitic magma had transported a certain amount of more basic magmas, judging from the wide occurrence of basic and intermediate inclusions in the lavas of that eruption.

As a result of deep degassing and the discharge of large magma volumes, the driving mechanism of the eruptive process had been exhausted for the time being, so that the volcano was in a state of repose for several decades. During that period the shallow chamber contained no free volatiles (the gas phase), but the amount of water dissolved in its melt was in excess of its equilibrium percentage. It was this water excess that served as a hidden driving force behind the moderate explosive activity observed in 1907–1931. The water might be released by two processes: (1) upward diffusion, (2) convective motion of deep-seated magma portions toward the top of the chamber. The latter process may have been triggered by decreasing melt density as water was being dissolved in it. The explosive process did not manifest itself at once, either owing to the low diffusion rate or to the moderate rate of convection, and was extended in time, with the vigor of the process clearly increasing over time. This phenomenon was especially noticeable a few years before the Plinian eruption during the resumed activity of Quizapu in 1926–1929 as described in [12].

Nevertheless, the time interval between 1847 and 1932 was probably insufficient to get rid of much of the excess water by diffusion. Consequently, the magma in the shallow chamber was already unable in 1932 to sufficiently accumulate the volatiles that were released from a new portion of water-oversaturated dacitic magma, so that the last eruption was Plinian in character. The basic magmas in the shallow chamber were richer in dissolved volatiles in 1932 compared with 1846, as shown by the copious basic cinder discharged by the Plinian eruption as compared with the fine-porous mafic inclusions in the lavas discharged by the first eruption. The causes that again set in motion the dacitic magma from the deeper chamber are not clear, neither is it clear why this magma portion was not discharged by the first eruption. The new portion was obviously rising from the same depth as the preceding one. This is shown both by the identical chemistry and mineralogy of the associated rocks and by the excellent preservation of hornblende. The last circumstance was also favored by the fact that the ascent rate of the 1932 dacitic magma seems to have been even higher than that in 1846, since the magma had kept much of the gas phase, was moving along a preexistent channel, and was discharged through a preexistent vent, which had been highly active previously. Cerro Azul Volcano, the Quizapu cone, Descabezado Volcano, and the new vent on its flank are all on a single line. This indicates that the 1932 magmatic and volcanic activity was occurring along a fault zone that had involved a sufficiently large area and had enhanced the magma permeability of the basement.

## DISCUSSION

The chief point of the mechanism proposed here is the assumption of two relatively independently evolving magma chambers, and that the source of dacitic magma must lie deeper than that of the more basic magmas. In our opinion, a deeper position of silicic magma relative to the basic magma is quite a feasible assumption with the existence of two spatially isolated chambers, and cannot be viewed as something unacceptable. For the case of the activity of Quizapu this structure of the feeding system permits all observed facts to fit nicely into a logical scheme.The complex structure of the feeding zone for Cerro Azul and Quizapu is corroborated by the polymodal composition of the 1932 ejecta.

Analogous inferences relating to the structure of feeding magma systems were drawn in [9, 11]. Doubic [9] analyzed a set of petrologic and volcanological data for the eruptions of the Shtyubel' cone (the center of recent activity on Ksudach Volcano, Kamchatka). One characteristic feature of the 1907 Plinian-type eruption on Ksudach was that its ejecta were inhomogeneous: at the beginning of the eruption the ejecta consisted of basaltic andesite to andesite cinder followed by a sharp change in the discharged rock composition, dacitic pumice being ejected with typical banded varieties. The author of this reference came to the conclusion that the Shtyubel' feeding system consists of two magma reservoirs, a shallow low-pressure chamber with its top probably at a depth of ~3 km and a deeper chamber of higher pressure. The shallow chamber is filled with basaltic andesite magma, while primary high-alumina basaltic magma is evolving toward dacite in the deeper chamber. Major explosive eruptions involved injection of dacitic magma from the deeper chamber into the basaltic andesite magma of the shallow chamber. It was the decompression of a silicic magma, rich in water and colder, that supposedly triggered the simultaneous eruption of two magmas of contrasting compositions. The simultaneous supply of a dacitic and a basaltic andesite magma led to the mixing of these in the volcanic conduit and to the generation of banded lavas that are typical of the Shtyubel' explosive eruptions. During the repose period after a major explosive eruption the two magmas that remain in the channel will mix to produce andesites; these possess a number of characteristics that reveal their hybrid origin. Here, as well as for the Quizapu case, the observed facts can be accounted

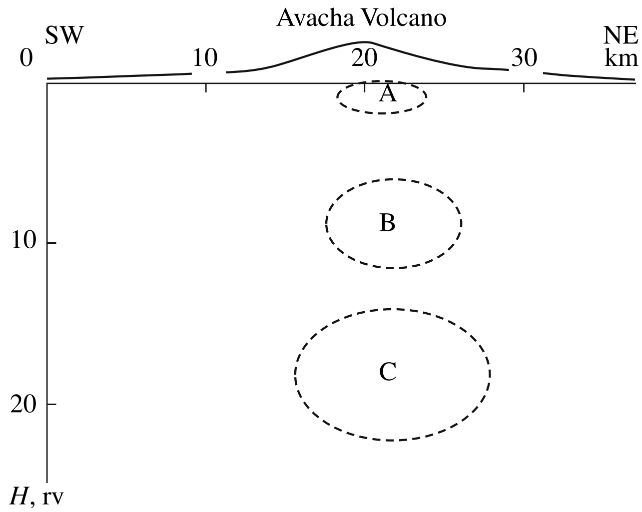


**Fig. 2.** Schematic representation of magma chambers beneath Avacha Volcano after [5]: (A) peripheral magma chamber, (B) intrusive body, (C) hypothetical crustal magma chamber.

for by interaction of magmas from two chambers at different depths, with the silicic magma being deeper.

Eichelberger and Izbekov [11] examined the 1912 eruption of the Katmai–Novarupta pair, with the eruption being one of the largest to occur during the 20th century. According to the above mechanism, the eruption can be explained by interaction between rhyolite and andesite magmas, with the silicic magma being originally at a greater depth. The rhyolite magma was rising along a dike that partially opened a shallow chamber containing andesite magma. The result was that the andesite magma flowed from this chamber toward the new eruptive center (Novarupta) and underwent partial mixing there with the rhyolite magma, producing dacitic magma. As a result, a large volume of dacite and subordinate amounts of andesites were erupted, in addition to the prevailing rhyolites. Both eruptions of the pairs Katmai–Novarupta (1912) and Cerro Azul–Quizapu (1846–1847) were the largest in the respective continents in their volumes of ejected magma, and were the first eruptions at these centers during historical times. It is a noteworthy fact that the dominant types of erupted rocks had been absent from these regions previously. While highly acidic rocks (rhyolite) were absent in the Katmai area, rocks of similar acidity (acid dacite) had previously been erupted in the Cerro Azul area, with the new factor being the appearance of hornblende in the rocks.

Hildreth and Drake [12] also noted a very great similarity between the activities of Quizapu and Mount Mazama in the Cascade Range, as well as between the general volcanisms of these areas. In particular, the paroxysmal Plinian-type eruption of Mount Mazama was preceded by discharges of large volumes of hornblende rhyodacite lavas, which contained numerous mafic inclusions, with the magma compositions of the effusive and the Plinian-type eruption being identical.

That the conception put forward here is a viable mechanism is indirectly corroborated by geophysical data on the structure of the feeding system beneath Avacha Volcano (Fig. 2) [5]. This data set implies the existence of two magma chambers at different depths beneath the volcano. The top of the shallower chamber is under the volcanic cone at depths of 0–2 km, while the deeper chamber is at the middle/lower crustal levels (of the order of 15–25 km). An intrusive body has been identified between the two chambers, which reminds one of the position of the granodiorite massif at the base of Cerro Azul Volcano. The initial phase of activity for the Molodoi Cone of Avacha Volcano typically involved Plinian-type eruptions of hornblende andesite with subsequent transition to eruptions of basaltic andesite ejecta [1]. Eruptions took place during this transition period whose ejecta included both of these compositions, which were first dominated by andesites and afterwards by basaltic andesites. The contrasting compositions of these ejecta also seem to have been a consequence of the fact that the material was coming from different chambers.

The interaction of magmas that have different volatile concentrations disturbs the equilibrium distribution of volatiles in a chamber. This may cause a long-continued degassing that is seen at the surface as long-continued fumarole activity and weak to moderate explosive activity. The latter is frequently interpreted as phreatic, but since it is not due to the interaction of magma and the water of the host rocks, this is essentially a kind of magmatic activity.

## CONCLUSIONS

Certain features in the vertical distribution of water (and other volatiles) in the melt can make a magma chamber a trap for the volatiles that are released from a portion of deep magma coming into it. This model of deep degassing can explain effusive discharges of originally aqueous magmas, including silicic magmas. After the lapse of several decades or hundreds of years an effusive eruption (or phase of activity) may be followed by a large Plinian-type eruption due to the interaction of magmas from chambers at different depths when the shallower chamber is no longer in a condition to accomodate a large mass of volatiles from the deep aqueous magma. This seems to be one of the most effective mechanisms for causing catastrophic eruptions.

The appearance of several contrasting compositions during a single eruption is not necessarily due to the zonality of a single magma chamber. On the contrary, this may point to different depths of the magma chambers. It should be admitted that the generally accepted viewpoint that the chambers of silicic magma are shallower relative to those of basic magma is not universal.

Such complex-structured magma systems may have been generated by tectonomagmatic activation subsequent to long repose periods during which the magma was undergoing deep differentiation.

## ACKNOWLEDGMENTS

This work was supported by the Russian Foundation for Basic Research, project no. 05-05-65300.

## REFERENCES

1. Bazanova, L.I., Braitseva, O.A., Puzankov, M.Yu., and Sulerzhitskii, L.D., Catastrophic Plinian Eruptions during the Initial Phase in the Generation of a Young Cone on Avacha Volcano, Kamchatka, *Vulkanol. Seismol.*, 2003, no. 5, pp. 20–40.
2. Kadik, A.A., Lebedev, E.B., and Khitarov, N.I., *Voda v magmaticheskikh rasplavakh* (Water in Magma Melts), Moscow: Nauka, 1971.
3. Kadik, A.A., Maksimov, A.P., and Ivanov, B.V., *Fiziko-khimicheskie usloviya kristallizatsii i genezis andezitov (na primere Klyuchevskoi gruppy vulkanov)* (Physicochemical Conditions of Crystallization and the Origin of Andesites: The Klyuchevskoi Group of Volcanoes), Moscow: Nauka, 1986.
4. Kennedy, G., Some Aspects of the Role of Water in Rock Melts, *Geol. Soc. Am. Spec. Paper*, 62, 1955, pp. 489–504.
5. Moroz, Yu.F. and Gontovaya, L.I., The Deep Structure of the Avacha–Koryakskii Volcanic Group, Kamchatka, *Vulkanol. Seismol.*, 2003, no. 4, pp. 3–10.
6. Ponomareva, V.V. and Tsyurupa, A.A., On Long Flows of Acid Liquid Magma on Krashenninikova Volcano, *Vulkanol. Seismol.*, 1985, no. 3, pp. 85–92.
7. Yaroshevskii, A.A., The Physicochemical Principles for the Behavior of a Melt + Gas Phase Magma System in the Gravity Field, *Geokhimiya*, 2003, no. 5, pp. 561–569.
8. Burnham, C.W., Development of the Burnham Model for Prediction of the $H_2O$ Solubility in Magmas, *Rev. Mineralogy*, 1994, vol. 30, pp. 123–129.
9. Doubic, Ph., *Study of the Eruptive Products from the Three Volcanoes in Kamchatka: Insights into Shallow Magma Evolution and Dynamics*, Ph. D. Thesis, SUNY at Buffalo, USA, 1997.
10. Eichelberger, J.C., Carrigan, C.R., Westrich, H.R., and Price, R.H., Non-Explosive Silicic Volcanism, *Nature*, 1986, vol. 323, pp. 598–602.
11. Eichelberger, J.C. and Izbekov, P.E., Eruption of Andesite Triggered by Dike Injection: Contrasting Cases at Karymsky Volcano, Kamchatka and Mount Katmai, Alaska, *Phil. Trans. R. Soc. Lond. A*, 2000, vol. 358, pp. 1465–1485.
12. Hildreth, W. and Drake, R.E., Volcan Quizapu, Chilean Andes, *Bull. Volcanol.*, 1992, vol. 54, no. 2, pp. 93–125.
13. Holloway, J.R. and Blank, J.G., Application of Experimental Results to C–O–H Species in Natural Melts, *Rev. Mineralogy*, 1994, vol. 30, pp. 187–230.
14. Merzbacher, C. and Eggler, D.H., A Magmatic Geohydrometer: Application to Mount St. Helens and Other Dacitic Magmas, *Geology*, 1984, vol. 12, pp. 587–590.
15. Rutherford, M.J. and Devine, J.D., The May 18, 1980 Eruption of Mount St. Helens: 3. Stability and Chemistry of Amphibole in the Magma Chamber, *J. Geophys. Res.*, 1988, vol. 93, pp. 949–959.
16. Rutherford, M.J. and Hill, P.M., Magma Ascent Rates from Amphibole Breakdown: An Experimental Study Applied to the 1980–1986 Mount St. Helens Eruptions, *J. Geophys. Res.*, 1993, vol. 98, no. B11, pp. 19 667–19 685.
17. Rutherford, M.J., Sigurdsson, H., Carey, S., and Davis, A., The May 18, 1980 Eruption of Mount St. Helens: I. Melt Composition and Experimental Phase Equilibria, *J. Geophys. Res.*, 1985, vol. 90, pp. 2929–2947.
18. Scaillet, B. and Evans, B.W., The 15 June 1991 Eruption of Mount Pinatubo. 1. Phase Equilibria and Pre-Eruption P–T–$fO_2$–$fH_2O$ Conditions of the Dacite Magma, *J. Petrology*, 1999, vol. 40, no. 3, pp. 381–411.